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Observed controls on resilience of groundwater to climate variability in Africa

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Summary Paragraph

Groundwater in Africa supports livelihoods and poverty alleviation^{1,2}, maintains vital ecosystems, and strongly influences terrestrial water and energy budgets³. However, hydrologic processes governing groundwater recharge sustaining this resource, and their sensitivity to climatic variability, are poorly constrained^{4,5}. Here we show, through analysis of multi-decadal groundwater hydrographs across sub-Saharan Africa, how aridity controls the predominant recharge processes whereas local hydrogeology influences the type and sensitivity of precipitation-recharge relationships. Some humid locations show approximately linear precipitation-recharge relationships with small rainfall intensity exceedance thresholds governing recharge; others show surprisingly small variation in recharge across a wide range of annual precipitation. As aridity increases, precipitation thresholds governing initiation of recharge increase, recharge becomes more episodic, and focussed recharge via losses from ephemeral overland flows becomes increasingly dominant. Extreme annual recharge is commonly associated with intense rainfall and flooding events, themselves often driven by large-scale climate controls. Intense precipitation, even during lower precipitation years, produces substantial recharge in some dry subtropical locations, challenging the ‘high certainty’ consensus that drying climatic trends will decrease water resources in such regions⁴. The likely resilience of groundwater in many areas revealed by improved understanding of precipitation-recharge

58 relationships is critical for informing reliable climate change impact projections and adaptation
59 strategies.
60

61 **Main Text (1983 plus 347 in figures)**
62 Groundwater is a fundamental component of the global hydro-climatic system^{3,5} and plays a central
63 role in sustaining water supplies and livelihoods in sub-Saharan Africa due to its widespread
64 availability⁶, generally high quality, and intrinsic ability to buffer⁷ the impacts of episodic drought
65 and pronounced climate variability that characterizes this region¹. Groundwater in sub-Saharan Africa
66 is poised to enable increased freshwater withdrawals as demand rises⁸ and climate change increases
67 variability in surface water resources. It is therefore critical to understand the renewability of
68 groundwater under current and future climate. Groundwater levels and fluxes are governed by a
69 dynamic interplay between recharge (replenishment of groundwater) and discharge (loss of
70 groundwater to streams, lakes, oceans or atmosphere) with a variety of controls and feedbacks from
71 climate, soils, geology, landcover and human abstraction⁹. It is notoriously difficult¹⁰ to determine
72 variations in recharge magnitudes over time and space and their relationship to climate as direct, long-
73 term observations of groundwater levels to inform such understanding in this region are sparse¹¹.
74 Regional water security assessments have therefore relied heavily on large-scale hydrological models
75 to derive estimates of potential groundwater resources across the continent⁸ but these remain
76 unvalidated by groundwater observations^{5,12}. A robust, data-driven, understanding of groundwater
77 recharge, and critically its dependence on climate, is fundamentally required to inform water resource
78 decision-making. Improved understanding of groundwater-climate sensitivity is also integral to
79 understanding important hydro-climate-ecological-human interactions across the region, both in the
80 present day¹³ and the deeper past¹⁴.
81
82 We address this challenge here by exploring precipitation-recharge (P-R) relationships across a
83 diverse range of climatic and geological contexts in sub-Saharan Africa, using a unique archive of
84 multi-decadal, groundwater level hydrographs (time series). By applying a consistent methodology
85 across the archive we are able to characterize the climate-groundwater relations observed into

86 indicative types which each lead to implications for understanding climate change impacts on
 87 groundwater systems and sustainable water management.

88

89 We contend that long term (i.e. decadal or longer) groundwater level hydrographs, with little or
 90 known interference from human activities, offer the most direct way of assessing variations in
 91 groundwater storage and, via inversion using a water table fluctuation (WTF) technique (see
 92 Methods), assessing temporal sensitivity of groundwater recharge to climate variability. We have,
 93 therefore, collated new unpublished records and updated previously published records to evaluate
 94 recharge and relationships with climate using a WTF methodology. The 14 multi-decadal
 95 hydrographs and accompanying precipitation records collated from nine countries in Sub-Saharan
 96 Africa cover a wide range of climate zones from hyper-arid to humid, including both the unimodal
 97 precipitation regimes (local summer wet season) of the northern and southern hemisphere subtropics
 98 and bimodal Equatorial regime, as well as a diverse range of geological and landscape settings
 99 (Figure 1, Table S1).

100

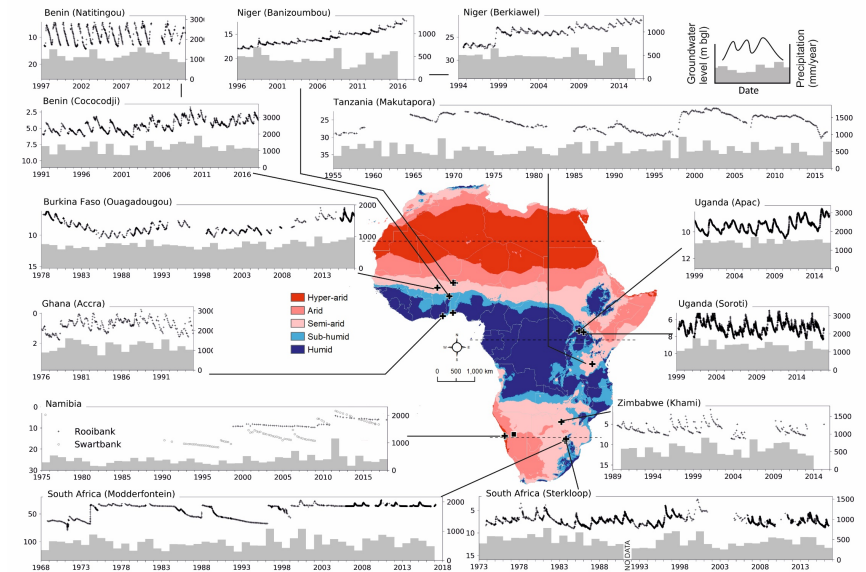
101 Most groundwater hydrographs show seasonal groundwater-level rises of varying magnitude that
 102 indicate recharge in excess of net groundwater drainage at some point during most years on record.
 103 The exceptions are Tanzania, Namibia and South Africa (Modderfontein) where multi-year
 104 continuous groundwater-level declines are observed, punctuated by episodic recharge events
 105 (Figure 1). Long term rising trends observed in the Niger hydrographs reflect increases in recharge
 106 rates since clearance of native vegetation in the 1960s¹⁵ which have not yet equilibrated with rates of
 107 net groundwater drainage due to long groundwater response times⁹ in the area. The absence of long
 108 term trends in other areas indicates a relatively stable balance between long term (i.e. multi-decadal)
 109 rates of groundwater recharge and discharge.

110

111 Groundwater recharge is often described on a continuum between ‘focussed’ (or ‘indirect’) recharge
 112 taking place via leakage of ephemeral streams or ponds, to ‘diffuse’ (or ‘direct’) recharge occurring
 113 in a more evenly distributed manner via the direct infiltration of precipitation at the land surface^{16,17}.
 114 The predominance of focussed recharge is thought to increase with aridity¹⁸ although there is no
 115 established threshold for when this occurs and diffuse recharge can also be significant in some semi-
 116 arid areas¹⁹. As part of conceptual models derived for each site, we developed a process-based
 117 understanding of recharge, resolving specifically whether diffuse or focussed recharge is dominant.
 118 This was assessed for each location based on additional reports, data, local knowledge and analysis
 119 of the form of the groundwater hydrographs themselves (see Methods, Table S3). We found that the
 120 transition from focussed-dominated to diffuse-dominated recharge occurs around the boundary
 121 between semi-arid and sub-humid conditions (Aridity Index, $P/PEt \approx 0.5$, Table S1).

122

123 **Figure 1. Collated multi-decadal groundwater-level and precipitation time series in sub-Saharan Africa.** A wide
 124 range of hydrograph responses, e.g. relatively consistent (Natitingou, Benin) or highly variable (Cococodji, Benin) annual
 125 fluctuations, highly episodic variations (Namibia, Tanzania), inter-decadal oscillations (Ouagadougou) or long term



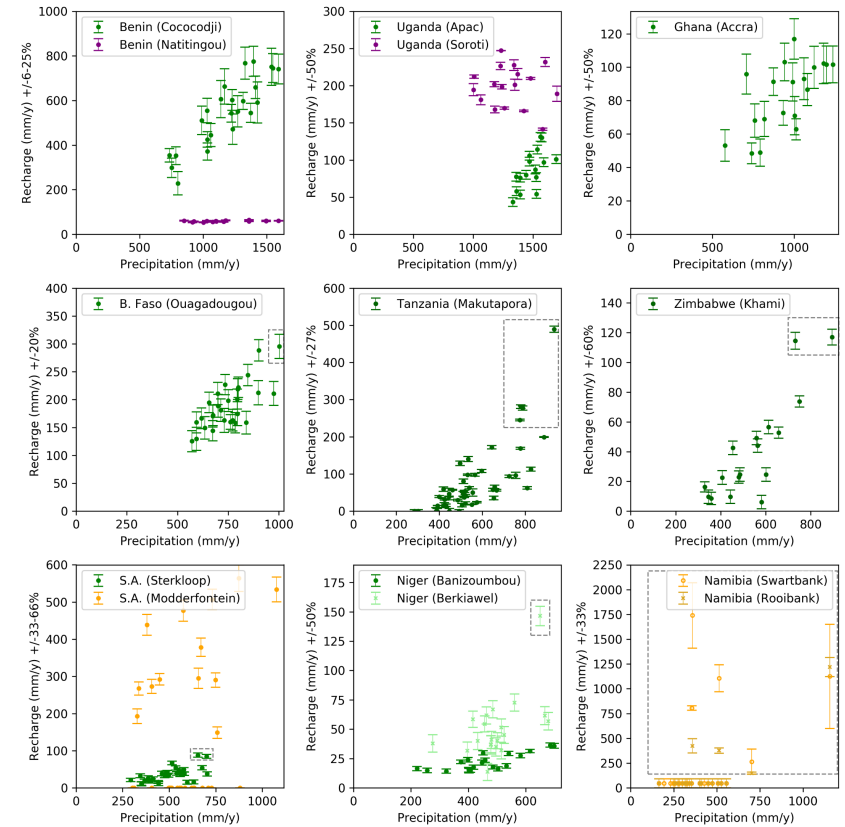
trends (Niger), reflect the complex interplay of climate, geology, soils and landcover represented across the monitoring locations. The analysed Namibian rain gauge is indicated by a filled black square. Aridity index classes are defined by the CGIAR-CSI Global-Aridity and Global-PET Database²⁰.

129

We have classified hydrographs according to their sensitivity of annual recharge to precipitation as reflected in the annual precipitation-recharge cross plots (herein ‘P-R plots’, Figure 2) and an analysis of how the proportion of recharge accumulates when years are ranked by annual precipitation (herein ‘rP-cR plots’, Figure 3, Figure S3). We then used a suite of idealised forward recharge modeling experiments to investigate how observed precipitation-recharge relationships relate to the magnitude of precipitation thresholds required to initiate recharge (see Methods and Figure S3). We observe three distinct types of precipitation-recharge sensitivity based on the empirical relationships derived from the data as follows (see Methods for site by site details):

(1) *Consistent recharge rates from year to year across the range of annual precipitation* (purple in Figures 2 to 4). This regime, exemplified by Natitingou (Benin) and Soroti (Uganda) shows little variation in annual recharge across a wide range of precipitation on P-R plots (Figure 2) and lies close to the 1:1 line in rP-cR plots (Figure 3). This type of precipitation–recharge response is found in sub-humid to humid locations and reflects the impact of local geology and soils in governing diffuse recharge processes.

144



145

146 **Figure 2. Observed relationships between precipitation and groundwater recharge on an annual (hydrological**
 147 **years) basis (P-R plots).** Error bars are based solely on uncertainty in recession within the WTF method. A best estimate
 148 of specific yield was used to estimate groundwater recharge values. Percentage errors in recharge due to uncertainty in
 149 specific yield as stated on the y-axis will result in a linear rescaling of values along that axis, but not alter the form of the
 150 relationships. Dashed boxes outline Tukey outlier values of extreme recharge. Note variable axis ranges. Sites are colour
 151 coded to represent the precipitation-recharge relationship types defined in Figure 3 and also used in Figure 4.

152

153

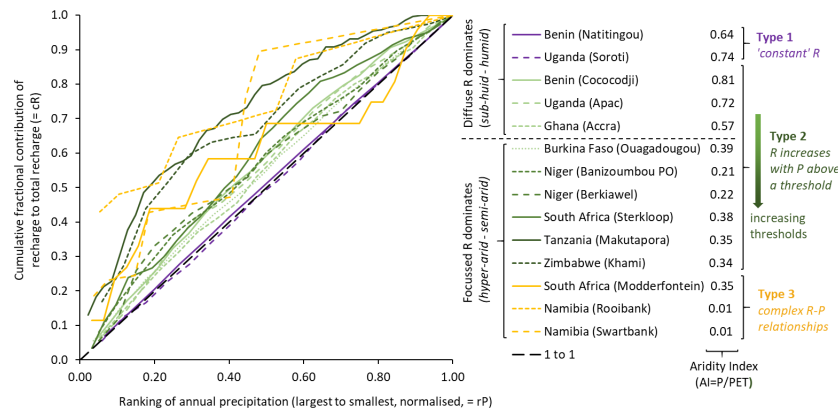


Figure 3. Cumulative contribution of annual recharge (by hydrological year) to total recharge for ranked annual precipitation (largest to smallest) (rP-cR plots). Values on both axes have been normalized by the total number of years in the record to provide fractional contributions for comparative purposes. Categorisation as either predominantly diffuse or focussed recharge is made on the basis of derived site conceptual models described in Table S3. Site colour coding is consistent with Figures 2 & 4.

(2) Increasing annual recharge with annual precipitation above a threshold (green in Figures 2 to 4). This type of regime shows positive P-R correlations (Figure 2) and shifts in the rP-cR relationship increasingly deviating to the left of the 1:1 line (Figure 3). This type is found at a majority of sites (n=9), across a wide range of aridity from humid to semi-arid conditions, and in areas dominated by both diffuse or focussed recharge. Sites with the largest apparent precipitation thresholds for recharge are located in semi-arid regions (Tanzania, Zimbabwe and South Africa-Sterkloop).

(3) Complex relationships between annual precipitation and recharge amount (orange in Figures 2 to 4). This type shows greater scatter on the P-R plots (Figure 2) and large 'steps' in rP-cR plots (Figure 3) as shown by Swartbank and Rooibank (Namibia) and Modderfontein (South Africa). A key feature of the annual P-R relationship is that some of the largest recharge events can occur

during relatively low total precipitation years. This type is found in semi-arid to hyper-arid locations dominated by focussed recharge.

Key insights regarding the relationships among aridity, recharge frequency, dominant recharge process and rP-cR relationships across the records are synthesized in Figure 4. This indicates the complex reality of controls on groundwater recharge and a lack of one to one correspondence with any individual factor. For example, while there is some relationship between the rP-cR relationships and degree of aridity (Figure 3 and Figure S2d,e), variation in local conditions (principally in soils/geology and precipitation intensity) results in distinctive characteristics in each location's recharge response to precipitation (see Methods). Hence, as aridity increases, while there is a transition from seasonal to episodic recharge frequency and from diffuse to focused recharge, there is also a significant spread of rP-cR types across different climates. Whilst not informed directly by our data, we also recognize that groundwater in some currently hyper-arid regions was recharged when a wetter climatic regime prevailed in the past (referred to as having 'Paleo' recharge frequency in Fig 4).

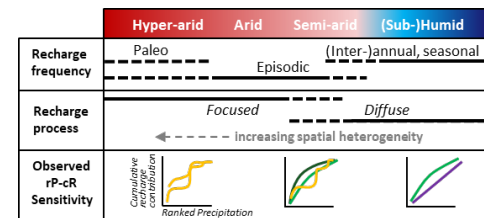


Figure 4. Idealised schematic of controls on recharge variations and processes in time and space in sub-Saharan Africa. As aridity increases, groundwater recharge tends to become increasingly heterogeneous in both space and time. Where recharge occurs via focused pathways, recharge may become 'increasingly indirect' as aridity increases meaning that the distance between the location of rainfall and the location of recharge increases. Paleo-recharge refers to recharge that occurred in some currently hyper-arid regions when a wetter climatic regime prevailed.

195

196 Where larger P thresholds for R are inferred, a smaller proportion of precipitation years yields the
 197 majority of long term recharge, and a majority of the variance in this relationship can be explained
 198 by increased aridity or coefficients of precipitation variability (Figure S2d-g), where wetter years
 199 contribute disproportionally to recharge. Further, values of extreme annual recharge identified as
 200 Tukey outliers (see Methods) were only found in more arid locations (AI<0.4, Table S1). By
 201 considering the wider regional precipitation distribution and the associated climate drivers during
 202 those years, we find that most years of substantial recharge are associated with widespread regional
 203 and seasonal scale precipitation anomalies, themselves associated with major known modes of global
 204 and/or regional climate variability (Table S2, Figure S4). As such, substantial variability in local
 205 groundwater recharge in piezometric records reflects the local impact of large-scale climate
 206 processes.

207

208 The different precipitation-recharge sensitivities observed have clear implications for understanding
 209 potential changes to groundwater levels and fluxes under climate change and therefore for developing
 210 sustainable strategies for groundwater provision for water supply or improving food security in Sub-
 211 Saharan Africa. Type 1 relationships imply that climate change impacts on precipitation may have
 212 little impact on recharge (other factors being equal). However, decreased groundwater levels due to
 213 pumping in such environments could provide more ‘room’ for recharge to occur via capture²⁶ of
 214 evapotranspiration (ET) or runoff. Increasing the distribution of groundwater monitoring in sub-
 215 Saharan Africa would help to identify Type 1 locations where groundwater abstraction can induce
 216 additional recharge. In these cases, and also for Type 2 sites with small P thresholds, sensitivity of
 217 recharge to changes in potential ET (PET) may also be low, because recharge is either not sensitive
 218 to P (Type 1) or factors other than P (Type 2) such as soil-moisture status. For Type 2 locations where
 219 thresholds are more highly influenced by antecedent dryness, recharge may be more sensitive to

220 climate change impacts on both precipitation and PET, and land use change could also be important
 221 if soil structure is altered and impacts runoff and infiltration processes²⁷.

222

223 The episodic nature of recharge in more arid locations and the preponderance of large groundwater
 224 response times⁹ in such areas together indicate the importance of long timescale planning horizons.
 225 In this context, the observed dependence of recharge on large-scale patterns of climate variability
 226 within Types 2 and 3 suggests the potential for a degree of predictability with seasonal lead times.
 227 Further it suggests that future changes in variability are likely to be of greater importance than mean
 228 precipitation. There is therefore a need to understand potential changes to such climate processes in
 229 longer multi-decadal climate change projections, currently a major challenge for climate models²⁸.

230

231 In contrast to rather uncertain recharge projections, modelled projections of increased flood hazards
 232 are more consistent for tropical Africa⁴ and focused recharge is likely to be widespread during such
 233 events. Hence, an important climate change adaptation strategy recommendation is for more
 234 widespread consideration of schemes to harness and enhance focused recharge during flood flow,
 235 storing water in the subsurface via managed aquifer recharge²⁹. Thus the increased flood risk under
 236 climate change may have a silver lining in this respect, water quality issues notwithstanding, and
 237 schemes to more effectively store flood water also have the potential to mitigate flood risk
 238 downstream. For Type 3, a key insight provided by our results in dry subtropical areas is that
 239 precipitation intensification, on the particular temporal and spatial scale determined by local
 240 conditions, may actually increase recharge, and thus available renewable water resources, despite an
 241 overall drying trend in annual precipitation totals³⁰.

242

243 Our data-driven results imply greater resilience to climate change than previously supposed in many
 244 locations from a groundwater perspective and thus question, for example, the model-driven IPCC
 245 consensus that “Climate change is projected to reduce renewable surface water and groundwater

resources significantly in most dry subtropical regions (robust evidence, high agreement)¹⁴. More observation-driven research is needed to clarify this issue, and address the balance of change between groundwater and surface water resources. Our results also pose a challenge to the reliance on standard large scale model assessments for inferring climate-groundwater dependencies until climate models can simulate with greater credibility both the large scale and local scale drivers of precipitation variability in the region, and hydrological models include the necessary recharge processes and the influence of geological variability. The establishment of greatly increased spatial coverage of long-term groundwater monitoring is needed to address the challenge of model validation in this context.

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281 **Methods (>4000 words – ie. a lot over the “ideally of no more than 3000 words”!)**

282 **Groundwater hydrograph and precipitation data collation and processing**

283 Multi-decadal time series of groundwater levels and precipitation were compiled by the authors from
284 records of observation wells initiated and maintained by government departments and research
285 institutions in nine countries in sub-Saharan Africa (Table S1, Figure 1). The pan-African collation
286 of these hydrographs was initiated at the 41st Congress of the International Association of
287 Hydrogeologists (IAH) in Marrakech (Morocco) on 14th September 2014. All records were subjected
288 to a rigorous review by the authors during which the integrity, continuity, duration and interpretability
289 of records were evaluated. This process included dedicated workshops in Benin, Tanzania, and
290 Uganda, and records failing these tests were discarded from the analysis. Procedures included taking
291 of the first time derivative to identify anomalous spikes in records commonly associated with errors
292 of data-entry. Where multiple records in same geographic and climate zone were available (e.g.
293 Benin, South Africa) we prioritized records remote from potential areas of intensive abstraction.
294 Statistical clustering of records was also used in the Limpopo Basin of South Africa to identify the
295 representativity of employed records at Modderfontein and Sterkloop. Hierarchical clustering was
296 done on hydrographs converted into a Standard Groundwater Index³¹ and identified three clusters
297 through a *kmeans* approach, one of which was an intermediary type hydrograph between two end
298 members represented by Modderfontein and Sterkloop.

299

300 Recognising that the substantial spatial variability of precipitation in sub-Saharan Africa may impact
301 observed relationships between precipitation and recharge, we used precipitation records which are
302 representative of the recharge generation process (i.e. diffuse or focussed). As a result, raingauges are
303 either co-located (e.g. < 5 m away) with groundwater monitoring sites or we employed the most
304 proximate rain gauge typically less than 10 km away (Table S1). In the case of the Namibian data,
305 the relevant raingauge was based more than 200 km away from the groundwater monitoring locations

306 to be representative of the runoff generation area in the Kuseb river, which acts as the source for
307 focussed recharge in these locations.

308

309 Each groundwater record thus has an accompanying daily (9 of the 14 records) or monthly (5 of the
310 14 records) precipitation record covering the same period. Infilling of occasional gaps of less than a
311 week in daily groundwater-level records was achieved by linear interpolation. All locations show
312 seasonal, mostly unimodal, precipitation (*P*) distributions with the exception of those in Uganda,
313 southern Benin (Natitingou) and Ghana with a more complex bimodal pattern (Figures S1).

314

315 **Relationships between average climatic variables, large scale climate processes and recharge**

316 Coefficients of monthly (or annual) precipitation variability were calculated as the standard deviation
317 of monthly (or annual) precipitation of the whole record divided by the mean precipitation of the
318 whole record multiplied by 100%. For analysis of wider climatic anomalies during major recharge
319 events we use gridded data of: Global Precipitation Climatology Centre (GPCC) monthly
320 precipitation product v8³² at 1.0° resolution; Daily precipitation at 0.1° resolution from the Climate
321 Hazards InfraRed Precipitation with Station Data³³; The Extended Reconstructed Sea Surface
322 Temperature (ERSST) version 4 data from the National Oceanographic and Atmospheric
323 Administration (NOAA)³⁴ on a monthly 2° grid.

324

325 Linear regression analysis indicates a strong correlation ($R^2=0.90$) between *P* and aridity index
326 (P/PEt) (Figure S2a) since rates of potential evapotranspiration (*PEt*) have a relatively small range
327 across these tropical latitudes in comparison to annual average rainfall. *PEt* neither correlates with *P*
328 ($R^2=0.00$) or P/PEt ($R^2=0.00$). Aridity index is strongly correlated to the coefficient of monthly *P*
329 variability ($R^2=0.82$), but less so with the coefficient of annual *P* variability ($R^2=0.36$) (Figure S2b,c)
330 together indicating that aridity is a strong control on the degree of rainfall seasonality.

331
 332 Long-term average recharge rates correlate poorly with rainfall or aridity (Figure S2h). In humid
 333 regions this is expected due to geological variations causing large differences in absolute recharge
 334 rates; in Benin for example, Cococodji recharge is nearly an order of magnitude greater than that in
 335 Natitingou despite similar rainfall and aridity (Figure 2). In more arid regions, increasing spatial
 336 heterogeneity in recharge rates is expected due to the increasingly predominance in focused recharge
 337 (Figure 4). Thus, the Namibian records, for example, show high rates of recharge reflective of the
 338 ‘footprint’ of the observation well located near an ephemeral stream; such values which are often
 339 higher than the local precipitation, would nevertheless be expected to be larger than average recharge
 340 rates for the wider hyper-arid region. Thus, the direct comparison of recharge rates between sites
 341 could be misleading without considering these potentially confounding factors.

342
 343 We show that most of the extreme recharge events, which are identified as recharge outliers (Figure
 344 2), are associated with relatively widespread regional and seasonal scale precipitation anomalies (see
 345 exemplar in Figure S3). These precipitation anomalies themselves can be associated with large-scale
 346 structures of climate variability known to impact the various regions of Africa (Table S2). Whilst
 347 recognising that observed precipitation variability typically results from a complex set of drivers
 348 occurring simultaneously over various spatial and temporal scales, we note the following association
 349 of large-scale precipitation anomalies during the outlier extreme local recharge years and climate
 350 drivers.

351
 352 Across our sites south of the equator we note that the major recharge years are associated with: El
 353 Niño events concurrent with the positive phase of the Indian Ocean Zonal Mode (IOZM³⁵) in the East
 354 Africa (Tanzania) site, and La Niña events in Southern Africa (South Africa and Zimbabwe, see
 355 Figure S3 as an example). This is consistent with the well-established north-south dipole precipitation
 356 response to El Niño-Southern Oscillation (ENSO) events which typically, but neither exclusively nor

357 consistently, bring wet (dry) rainfall anomalies across East (Southern) Africa during El Niño events
 358 and the reverse during La Niña^{36,37}.

359
 360 Further west in the hyper-arid Namibia sites the drivers of the outlier recharge events are more
 361 complex, as expected given the complex ‘Type 3’ relationship of precipitation to the highly episodic
 362 recharge (Figure 2), dependent on triggering of ephemeral surface river flow. Of the five outlier
 363 recharge events, two can be linked to regional/seasonal scale rainfall anomalies associated with an
 364 anomalous warming of the cold Benguela current off the west coast of Africa. Such ‘Benguela Niño’
 365 events³⁸ are known to trigger convection and rainfall across much of Northern Namibia and Southern
 366 Angola^{39,40}. The remaining three events appear linked to spatially extensive but shorter duration
 367 heavy rainfall anomalies from sub-seasonal variability. This includes the notable, anomalous
 368 westward propagation of tropical cyclone Eline in February 2000 from the Indian Ocean basin to
 369 Namibia, which also caused widespread precipitation extremes across much of South-eastern Africa
 370 compounding existing La Niña related rainfall, as well as synoptic scale tropical low pressure systems
 371 in 2009.

372
 373 The West African sites show a smaller number of outlier recharge events. The 2012 event at Burkina
 374 Faso appears part of wider, regional and seasonal scale precipitation anomalies, in which the Sahel
 375 region as a whole experienced the strongest monsoon season since 1953, likely resulting from the
 376 combination of seasonal tropical Atlantic temperature anomalies⁴¹ and sub-seasonal variability from
 377 active phases of the Madden Julian Oscillation⁴². The 1998 recharge event in Niger coincided with
 378 far less spatially coherent seasonal anomalies and likely resulted from intensive sub-seasonal
 379 precipitation events.

380

381 **Site conceptual models**

For each hydrograph location, a conceptual hydrogeological model was formulated based on available data, literature, and site visits by the authors, as necessary (Table S3). These included an assessment of the main hydrogeological boundaries such as groundwater divides and perennial or ephemeral drainage features; the local context for factors which may influence recharge such as geology, soils, climate variables, groundwater abstraction and the thickness of the unsaturated zone; and estimations of aquifer storage and transmissivity. A particular focus was to develop an appreciation, based on the local context, of how ‘diffuse’ the recharge is likely to be spatially, or whether ‘focussed’ recharge is likely to be more significant in causing local variations in the magnitude of water table fluctuations. Of most importance for determining the predominance of diffuse versus focussed recharge is: the presence or absence of perennial versus ephemeral streams; co-incident timing of ephemeral or seasonal stream flows with water table responses; and the form of groundwater hydrographs with respect to the presence or absence of groundwater mounding as indicative of focussed recharge (see further details in the Section ‘Groundwater recessions’ below). The conceptual hydrogeological model development also enabled us to ensure that observed groundwater level changes are likely to be representative of water table fluctuations in an unconfined aquifer (i.e. vertical flow in the aquifer is insignificant and that poro-elastic or other ‘confined’ responses are negligible).

Recharge estimation using water table fluctuation (WTF) method

Approach and equations: inverse WTF models were used to infer the recharge timing and magnitude at the location of each hydrograph. The WTF technique is the most direct method of transient groundwater recharge estimation available and has very few embodied assumptions in comparison to other methods such as geochemical tracers or modelling approaches^{16,43}. In a recent review of recharge estimation methods it was strongly recommended for application in humid and semi-arid African regions¹⁰ and it is also applicable for both diffuse^{44,45} and focussed⁴⁶ recharge situations.

We assume that groundwater level (or hydraulic head, h [L]) at an observation point is naturally controlled by the combined influence of the rate of net groundwater recharge (R [LT⁻¹]), balanced by the rate of ‘net groundwater drainage’ (D [LT⁻¹]) acting at that point in space and time. Further variations in WTF may be superimposed due to the rate of ‘net drawdown’ (s [LT⁻¹]) caused by changes in groundwater abstraction occurring at some distance from the observation point.

The following water balance equation was used to approximate a time series (with time step Δt) of the ratio of recharge (R_t) to specific yield (S_y [-]):

$$\frac{R_t}{S_y} = \frac{(h_t - h_{t-\Delta t})}{\Delta t} + GWL_r + s_t \quad (1)$$

where $GWL_r (= \frac{D_t}{S_y})$ is the rate of groundwater level recession⁴⁷ [LT⁻¹]. Values for absolute value of recharge were then also calculated by multiplying by the applicable specific yield at the position of the water table.

To enable exact accounting periods for comparison with precipitation records, and between hydrographs, where observations were less frequent than daily, linear interpolation was used between groundwater level observations. Calculations were carried out on a daily time step and sums were calculated for hydrological years (in both R and P). If observations were missing across either end of the hydrological year in the first or last years of record, those years were removed from further analysis. Within the annual recharge time series generated “Tukey” outliers were identified as any years with values greater than 1.5 times the interquartile range above the third quartile.

427 **Groundwater recessions:** the GWL_r term was estimated based on the observed form and magnitude
 428 of the groundwater hydrograph during long dry periods by either setting a constant rate, or an
 429 exponential decay controlled by the following equation:

$$430 \quad \quad \quad GWL_r = \frac{h(t-\Delta t) - h_b}{C} \quad (2)$$

431 Where h_b is the elevation of the assumed lateral groundwater drainage boundary [L] and C is a decay
 432 constant [T^{-1}].

433

434 Most of the hydrographs have very distinctive seasonal precipitation patterns with long dry seasons
 435 during which the true form of groundwater recession (i.e. a groundwater level decline in the absence
 436 of any recharge) can be directly observed, assuming no human interferences⁴⁷. This enables the choice
 437 of recession model (constant rate or exponential) to be confidently made, and constant rates or decay
 438 constants to be easily determined. This is in comparison to more humid parts of the world with limited
 439 dry periods where the WTF is harder to apply robustly^{44,45}. For hydrographs in Ghana, South Africa
 440 (Modderfontein) and Burkina Faso, an exponential recession model was used due to the presence of
 441 a shallow water table, inferred high permeability fracture flow, and close proximity of the
 442 groundwater drainage boundary, respectively (see Table S3 for more information). For Uganda
 443 (Soroti), the absence of long dry seasons, and the observed form of groundwater level declines did
 444 not make the choice between exponential and straight line recessions obvious, and so both were
 445 applied to represent the uncertainty. For all other locations, the observed variation of dry season
 446 groundwater-level recessions was used to define maximum and minimum constant rate end members
 447 to constrain the uncertainty in recharge estimates due to this parameter. This is consistent with
 448 theoretical expectations of linear recessions for these locations with drainage boundaries (where
 449 known) being sufficiently distant given the aquifer properties⁴⁷.

450

451 For the cases where focussed recharge is significant due to local infiltration from ephemeral streams
 452 and ponds, the expected theoretical form⁴⁶ is for groundwater hydrographs to show steep recessions
 453 following a rise in the water table, then trend to a relatively constant lower ‘background’ recession
 454 rate. This is observed for example in Zimbabwe, Tanzania, South Africa (Sterkloop), Niger and
 455 Namibia and is explained by localised groundwater ‘mounding’ near the location of focussed recharge
 456 dissipating on a much quicker timescale than the regional background recession, which operates on
 457 much longer spatial scales. In these cases, with the exception of Namibia, the local mounding
 458 dissipates before the end of the hydrological year enabling a seasonal WTF accounting period to be
 459 used following the method of ref⁴⁶. In Namibia, the recession of the recharge mounds occurred over
 460 timescales greater than a single season, and therefore had to be extrapolated leading to much greater
 461 uncertainty in the output recharge values (as evidenced by larger error bars in Figure 2). The
 462 application of this method thus enables recharge rates to be derived which are representative of
 463 integrated processes across larger areas of the catchment or aquifer (whichever define the hydraulic
 464 boundaries), rather than simply reflecting the local conditions near the stream. However, the spatial
 465 representativity of recharge estimated at each location is variable and, as such, direct comparisons of
 466 absolute recharge rates from site to site should only be made where this can be accounted for.

467

468 **Groundwater abstractions:** once at steady state, groundwater abstractions should have no effect on
 469 water table fluctuations. However, transient abstractions cause time-varying drawdown at the
 470 groundwater monitoring location. If not accounted for, they will therefore cause recharge
 471 underestimations when drawdown is increasing, and overestimations when drawdowns are
 472 decreasing (e.g. if abstraction temporally reduces (increases) causing recovery (decline) in
 473 groundwater levels). In most cases, the observation wells are located far from the influence of major
 474 changes in groundwater abstraction as documented in the meta-data (Table S3) and s_i was assumed
 475 to be zero. In one location (Makutapora, Tanzania), the monitoring wells are located within a major
 476 well-field where abstraction rates have been highly variable during the monitoring period. Corrections

22

were therefore made for this site using a 3D groundwater model to estimate a time series of net drawdown to account for the changes in recession due to variations in pumping rate. (i.e. accounting for drawdown due to increases in pumping, and recovery during decreases in pumping) (meta-data Table S3).

Specific yield: ranges of specific yield were estimated based on local information and literature for each groundwater level record as described in the meta-data (Table S3) and assumed to be constant in time, and across the range of water table fluctuations at a given location. The uncertainty in specific yield can be considerable and represents the main uncertainty in the derived absolute values of recharge. As well as being notoriously hard to estimate⁴⁸, it is also known that specific yield can vary in time due to vertical heterogeneity in lithology, due to shallow water tables or where swelling clays are present^{44,49}. The variation in the value for specific yield has no impact on the form of the relationship that recharge has with precipitation or the ranking of recharge events used in Figure 2 and 3 (and Figure S3). However, we report the likely range of uncertainty in specific yield for each location as this does impact the absolute magnitude of the recharge estimates, and is one reason why inter-site comparisons of long term average recharge by this method can be problematic.

Model experiments and interpretation of observed precipitation-recharge (rP-cR) relationships

P-R cross plots (Figure 2) showing annual recharge against annual precipitation, allow an initial characterization of precipitation-recharge relationships to be developed. For comparative purposes across all records, we then normalized annual recharge by a cumulative sum as a fraction of the total recharge for all years in a given record, and plotted this against the fractional precipitation ranking for each record (rP-cR plots, Figure 3). To inform process-based inferences from these plots, we ran a suite of numerical recharge model experiments using models with different structures, for two chosen time series from contrasting climates in sub-Saharan Africa: Dodoma (semi-arid Tanzania)

and Cococodji (humid Benin). The purpose was not to calibrate models for each of the locations across Africa but rather to understand the generic features of rP-cR plots for aiding interpretation of the relationships we observe in the data.

Three model structures of increasing complexity were explored: (a) Recharge was assumed to be constant for precipitation events above a daily or annual threshold. Note that, since the values were normalized against the total recharge in all years, the actual recharge value is irrelevant to the result. (b) The second model structure, in the manner of ref²¹, assumes that a constant proportion of precipitation becomes recharge above a specified daily precipitation threshold. Thresholds were applied at a daily time step and then results aggregated for yearly comparisons. Since the values were normalized against the total recharge in all years, the chosen proportion of rainfall that becomes recharge is irrelevant to the result. (c) The third model was a dynamic single layer soil moisture balance model (SMBM), in the manner of refs^{23,50}, also run at a daily time step and then aggregated to annual values. It was assumed in all SMBM model runs that the readily available water (RAW) was 50% of the total available water (TAW), that the crop coefficient was equal to 1 (e.g. for grass land cover), that the ratio of actual to potential evapotranspiration rates (AET/PET, a proxy for plant stress) decreased linearly from 1 to 0 as soil moisture deficit values increased from RAW to TAW, and that runoff was zero.

For Dodoma, daily PET values were derived using the Hargreaves and Samani equation⁵¹ from temperature data from the Dodoma Meteorological Station. In the case of missing data, the average value from the month is used or when, early in the record, entire months are without data the average temperature values for the corresponding month from the entire record was substituted. The calculated values were calibrated on pan evaporation data from the same location. For Cococodji,

526 daily PET values derive from pan evaporation data collected from the meteorological station at the
527 IITA (International Institute of Tropical Agriculture) office in Cotonou.

528

529 The generic style of each type of rP-cR plot (Figure 3) is well captured by the models, for either of
530 the two contrasting precipitation time series (Figure S3); both show three distinct types of
531 relationships and it is clear that different models (and thus processes) can lead to a similar sensitivity
532 – i.e. a critical point is that each type of observed P-R sensitivity does not necessarily correspond to
533 a particular recharge process. The first type (purple in Figure S3) plots close to the 1:1 line indicating
534 very consistent R values each year despite wide variations in P. The second type (green in Figure S3)
535 deviates from the 1:1 line increasingly as the size of the potential thresholds in the SMBM (governed
536 by TAW) or the actual thresholds in the linear models increase. The third type (orange in Figure S3)
537 shows pronounced steps in the curve generated by the largest thresholds in the linear model. Clearly,
538 P-R responses in reality fall on a continuum, but we propose that classifying by three types highlights
539 the end member responses. This classification can be further tested and refined as more data become
540 available for sub-Saharan Africa (or other parts of the world).

541

542 More details of the observed P-R and rP-cR plots (Figures 2 and 3) summarized in the main text are
543 as follows:

544 **Type 1:** Natitingou is characterised by low storage fractured bedrock ($S_y = 0.4\%$)²¹; water-table
545 variations are around 10 m annually and each year the subsurface fills to a shallow level. In
546 combination with straight recessions, this hydrogeological context leads to temporally small variation
547 of recharge each year despite large variations in annual precipitation (observed range is 850-1592
548 mm/y). At Soroti, the water table is always deeper than 5 m below ground level (bgl) within
549 weathered basement rock but, despite this, exhibits rapid responses to precipitation events indicative
550 of preferential flow processes²². The observed consistency in recharge from year to year may be

551 controlled by a finite near surface water store to which the water table responds²³ although further
552 site-investigations are needed to confirm precise controls.

553 **Type 2:** Where diffuse recharge is predominant, this type of sensitivity is expected if precipitation
554 thresholds are governed by prevailing soil moisture deficits (or other near-surface storage/losses). We
555 may expect increased deviation to the left of the 1:1 line on the rP-cR plots to increase with aridity
556 and the build-up of larger soil moisture deficits. However, we may also expect exceptions to this to
557 occur in cases where preferential flow processes²² are prevalent and recharge can ‘bypass’ soil
558 moisture deficits and thus precipitation thresholds may be lower than anticipated than under uniform
559 flow assumptions. Where focussed recharge is predominant, thresholds for its occurrence are
560 expected to be governed by hydrological processes which dominate in drylands, such as generation
561 of infiltration-excess runoff producing ephemeral channel flow²⁴. These processes can be locally
562 variable and have complex dependencies on, for example, land cover, drainage network density, soil
563 structure and antecedent moisture conditions. In the observed responses of this type in the humid to
564 sub-humid environments (i.e. Benin (Cococodji), Uganda (Apac) and Ghana (Accra)), thresholds
565 appear to be relatively small. This is consistent with detailed analysis available for Cococodji and
566 Apac which suggest values of 5 mm/d and <10 mm/d respectively for these sites; there are no existing
567 studies to corroborate this for Ghana. We observe that much larger precipitation thresholds may need
568 to be overcome for recharge to occur for semi-arid sites in Tanzania and Zimbabwe; here, the forward
569 models suggest precipitation thresholds greatly in excess of 100 mm/d (darker green in Figures
570 2 to 4). Again, this is consistent with detailed analysis carried out for Tanzania which indicates that
571 recharge occurs only after persistent rainfall of over 90 mm over a 7 day period⁵². For the two Niger
572 sites, despite also having greater aridity, thresholds are apparently much lower but this is explained
573 by daily precipitation thresholds of 10-20 mm d⁻¹ known to be required to generate stream flow²⁵, and
574 thus focussed recharge, in this area. In Burkina Faso, focussed recharge from a nearby managed
575 reservoir (“barrage”) is thought to moderate the impact of inter-annual precipitation variability on

576 recharge variability moving the rP-cR line closer to the 1:1 (Figure 3) than might be the case without
577 a reservoir.

578 **Type 3:** The two Namibian sites are in a hyper-arid environment dependent on runoff generation from
579 a large upstream catchment to supply focussed recharge during streamflow events. Conditions for
580 runoff generation are governed by intense monthly precipitation occurring not necessarily within
581 years of relatively high total precipitation. In contrast, at Modderfontein (South Africa), focussed
582 recharge is much more local, but the limestone bedrock in this location is typified by highly non-
583 linear hydrological processes which generate complex P-R relationships (see Table S3).

584

585 In summary, the controls on the observed P-R and rP-cR sensitivities are a complex interaction
586 between the prevailing climate and local controls on recharge generation.

587

588 **Data Availability Statement**

589 Where permissions allow, digital datasets are freely available to download online from
590 <https://doi.org/10.6084/m9.figshare.5103796> including calculated annual recharge values. and time
591 series of groundwater-level deviations from the mean and precipitation anomalies (also available at
592 IGRAC ‘link TBC’ and NERC ‘link TBC’).

593 **Author Contributions**

594 The paper was conceived by RT, GF and MOC. The paper was written by MOC and RT with input
595 from all authors. RT and GF led the collation of the data. MOC designed the applied WTF
596 methodology. MT conducted climate data analyses. All authors carried out other data analysis or
597 interpretation of results.

598

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